

CHAPTER D.2

PHYSICAL PROCESSES ALONG THE LOUISIANA COAST

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2.1 Summary

This chapter discusses the physical processes that are shaping coastal Louisiana. The major processes operating on the coast include waves, storms, tides, relative sea-level rise, tidal inlets and tidal prism, currents, and the movement of sediment. These processes are not inherently harmful. However, when combined with other man-made factors, such as levees and canals, many physical processes speed the destruction of Louisiana's coast.. Marsh and other wetlands are converting to open water, and barrier islands are disintegrating.

2.2 Introduction

The coastal barrier chains in Louisiana form the first line of defense for protecting wetlands, inland bays, and mainland regions from the direct effects of wind, waves, and storms. The barrier systems serve multiple purposes to:

- reduce coastal flooding during periods of storm surge;
- prevent direct ocean wave attack, which would accelerate rates of erosion and degradation of marshes and other wetlands; and
- help maintain gradients between saline and freshwater, thereby preserving estuarine systems.

The morphology and integrity of barrier islands along Louisiana's shoreline are directly related to the supply of sediment contributed to the coast and the physical processes operating in this region. The coastal zone is one of the most dynamic environments that exist in nature. The

same processes that built the barrier islands are also partly responsible for their erosion and fragmentation. This chapter discusses the coastal processes impacting the barrier islands, tidal inlets, and various backbarrier settings. A description of the waves, tides, circulation of coastal waters, tropical and extra-tropical storms, winds, tidal inlet dynamics, and subsidence and relative sea-level rise is presented in order to provide a basis for understanding the evolution of coastal Louisiana and assessing future possible changes to reconstructed barriers.

2.3 Waves

The wave climate along the Louisiana coast is a product of seasonal wind patterns and the passage of tropical and extra-tropical storms. The distribution of deepwater wave energy is known from several NOAA stations that are located 35 to 190 km (22 - 118 miles) offshore. The nearshore wave climate is less well known, and comes primarily from data sets that are collected at WAVCIS (Wave Current Information System) stations, an ocean observing system housed at Louisiana State University. Locations of the NOAA buoys and WAVCIS stations are shown in Figure D.2-1.

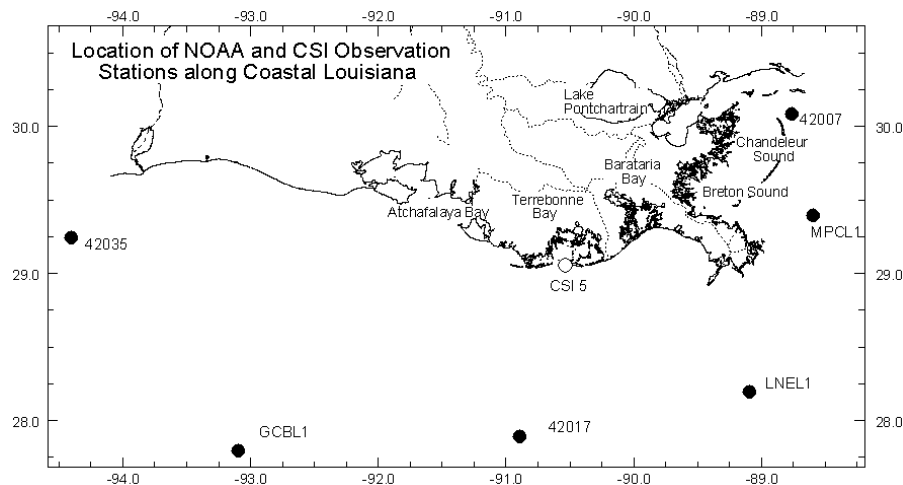


Figure D.2-1. Sites of deepwater wave gages (NOAA stations, solid circles and nearshore WAVCIS station (open circle).

2.3.1 Deepwater Waves

A summary of deepwater wave characteristics for the Louisiana coast is given Table D.2-1. The period of record ranges widely from 1 to 20 years. Despite this variation in temporal coverage, the trends are fairly consistent. The data indicate that the mean annual deepwater wave height varies from a low of 0.75 m (2.5 ft) off the central Louisiana coast (Station 42017) to a high of 0.95 m (3.2 ft) offshore of both Holly Beach (Station GCBL1) and Head of Passes (Station LNEL1). An exception to this trend occurs at Station 42007 northeast of the northern Chandeleur Islands where the average annual wave height is only 0.75 m (2.5 ft). The lower average wave height measured at this site as compared to other stations is due to the protection afforded by the Chandeleur Islands and Belize Delta from wave energy coming from the southwesterly quadrant.

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Wave periods at all the stations exhibit similar trends (Table D.2-1). Mean annual wave period ranges from 4.5 to 5.9 seconds, with the longest average period recorded at the station offshore of Holly Beach (GCBL1) and the shortest period measured at the northern Chandeleur Islands location (42007).

Table D.2-1. Summary of offshore wave data for the Louisiana coast.

Waves Table



Significant Wave Height Information

Station	Location	Offshore Distance (Km)	Water Depth (meters)	Average (meters)	Mean Range (meters)	Minimum (meters)	Maximum (meters)	Record Length
42035	Texas - West Louisiana	35	15.9	0.90	0.65 - 1.10	0.30	1.60	1993 - 2001
GCBL1	West (Holly Beach)	190	300	0.95	0.30 - 1.70	0.50	1.30	1990 - 1991
42017	Central (Terrebonne)	110	600	0.75	0.50 - 1.00	0.20	1.20	1989
LNEL1	Central (Head of Passes)	75	1,500	0.95	0.70 - 1.30	0.45	1.90	1991 - 1992
MPGL1	East (Breton Sound)	40	100	0.85	0.50 - 1.10	0.15	1.70	1988 - 1992
42007	East (Biloxi)	10	13.4	0.50	0.30 - 0.70	0.00	1.20	1981 - 2001

Periods Table

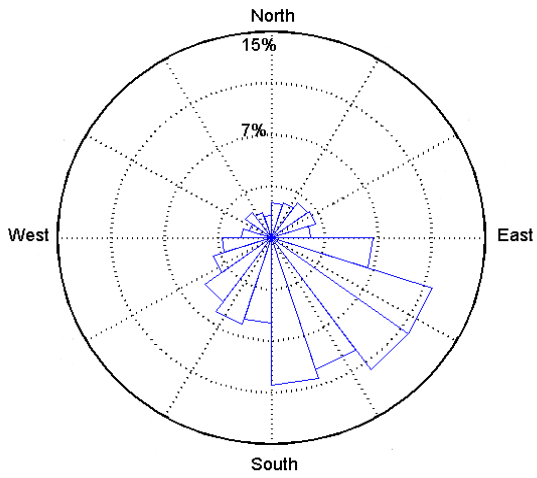
Dominant wave Period Information

Station	Location	Offshore Distance (Km)	Water Depth (meters)	Average (seconds)	Mean Range (seconds)	Minimum (seconds)	Maximum (seconds)	Record Length
42035	Texas - West Louisiana	35	15.9	5.50	5.00 - 6.00	3.80	7.50	1993 - 2001
GCBL1	West (Holly Beach)	190	300	5.90	4.70 - 6.70	3.50	8.00	1990 - 1991
42017	Central (Terrebonne)	110	600	5.00	4.90 - 5.10	3.00	7.00	1989
LNEL1	Central (Head of Passes)	75	1,500	4.65	4.50 - 5.20	3.80	6.10	1991 - 1992
MPGL1	East (Breton Sound)	40	100	5.50	5.00 - 6.00	3.00	7.00	1988 - 1992
42007	East (Biloxi)	10	13.4	4.50	4.00 - 5.00	2.00	7.00	1981 - 2001

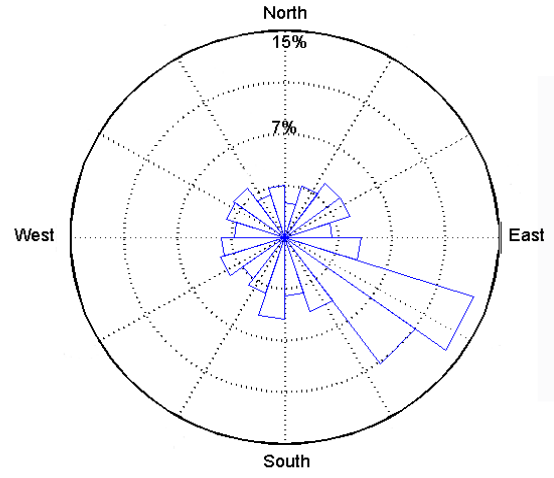
2.3.2 Nearshore Waves

The nearshore WAVCIS (CSI-5) station, which is located approximately 2 km off Timbalier Island at the 5 m (16.4 ft) isobath (Figure D.2-1), is particularly instructive in defining seasonal variations in wave climate for the central Louisiana coast (Figure D.2-2). The data in Figure D.2-2 demonstrate a high correspondence between dominant wave approach and wind direction. For the period between mid-spring and mid-fall, winds are predominantly from the south, with the highest frequency of occurrence from the southeast. Likewise, the dominant wave approach is from the southeast quadrant (40% probability). During the late fall to early spring, the wind regime is controlled by the passages of cold fronts (discussed in a later section). These weather systems commonly produce winds blowing from the south (pre-frontal) and then from the north (post frontal). However, the northeasterly winds blow offshore in central Louisiana, which cancels propagation of longer waves from the south near the coast. Fetch north of the WAVCIS stations is limited, and waves are not generated during these north events (Figure D.2-2). Therefore, the dominant waves (probability ~ 80%) come from the southeast quadrant and are the chief control of sediment transport patterns along the central Louisiana shore. Data from the same WAVCIS station (CSI-5), illustrates the seasonality in wave energy in Louisiana (Figure D.2-3). The graphs in Figure D.2-3 represent time series of wave height from May-August 2001 as measured by CSI-5. The lower graph shows the wave height distribution at the same location for a winter period extending from November 2001-February 2002.

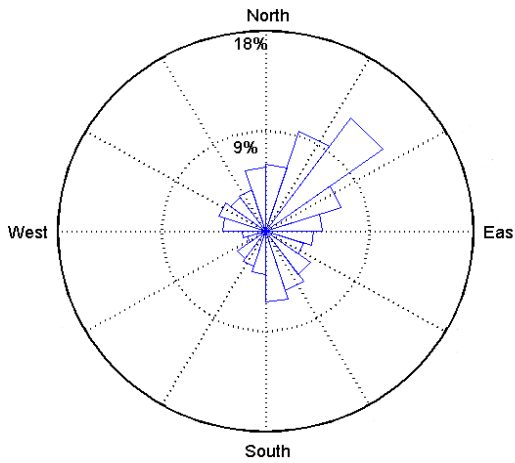
Typically, waves vary from approximately 0.07 m to 0.8 m (0.23 - 2.65 ft), the latter being a function of weak storms in the Gulf of Mexico. It is apparent, however, that tropical cyclones can generate considerably larger waves during summer months as shown here for a period in early June, 2001 where waves over 1.9 m (6.25 ft) in height were recorded. During winter months, the effects of cold front passages over the Louisiana coast are apparent as a series of sharp increases in wave height. During this four month period from November 2001 to February, 2002, 20 cold front passages can be identified with six events resulting in energetic sea states and wave heights ranging from ~1 -2 m (3.3 - 6.6 ft). Therefore, with the exception of infrequent tropical cyclone activity in summer months, the high frequency of frontal passages over the Louisiana coast plays a critical role in generating and sustaining higher waves during winter months.



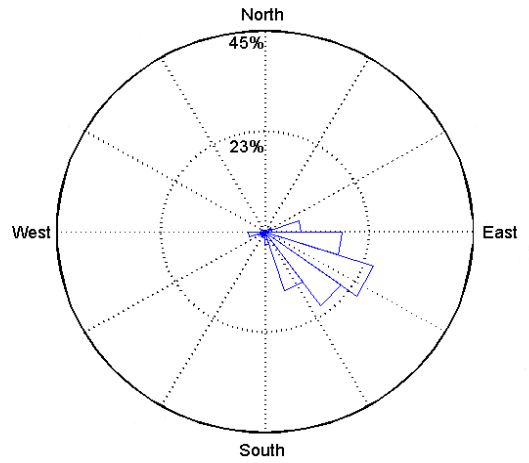
Wind Direction at CSI 5 (Summer)



Wave Direction at CSI 5 (Summer)



Wind Direction at CSI 5 (Winter)



Wave Direction at CSI 5 (Winter)

Figure D.2-2. Wave (right) and wind (left) roses for data collected at WAVCIS station (CSI-5) showing seasonal effects during summer (top) and winter (bottom).

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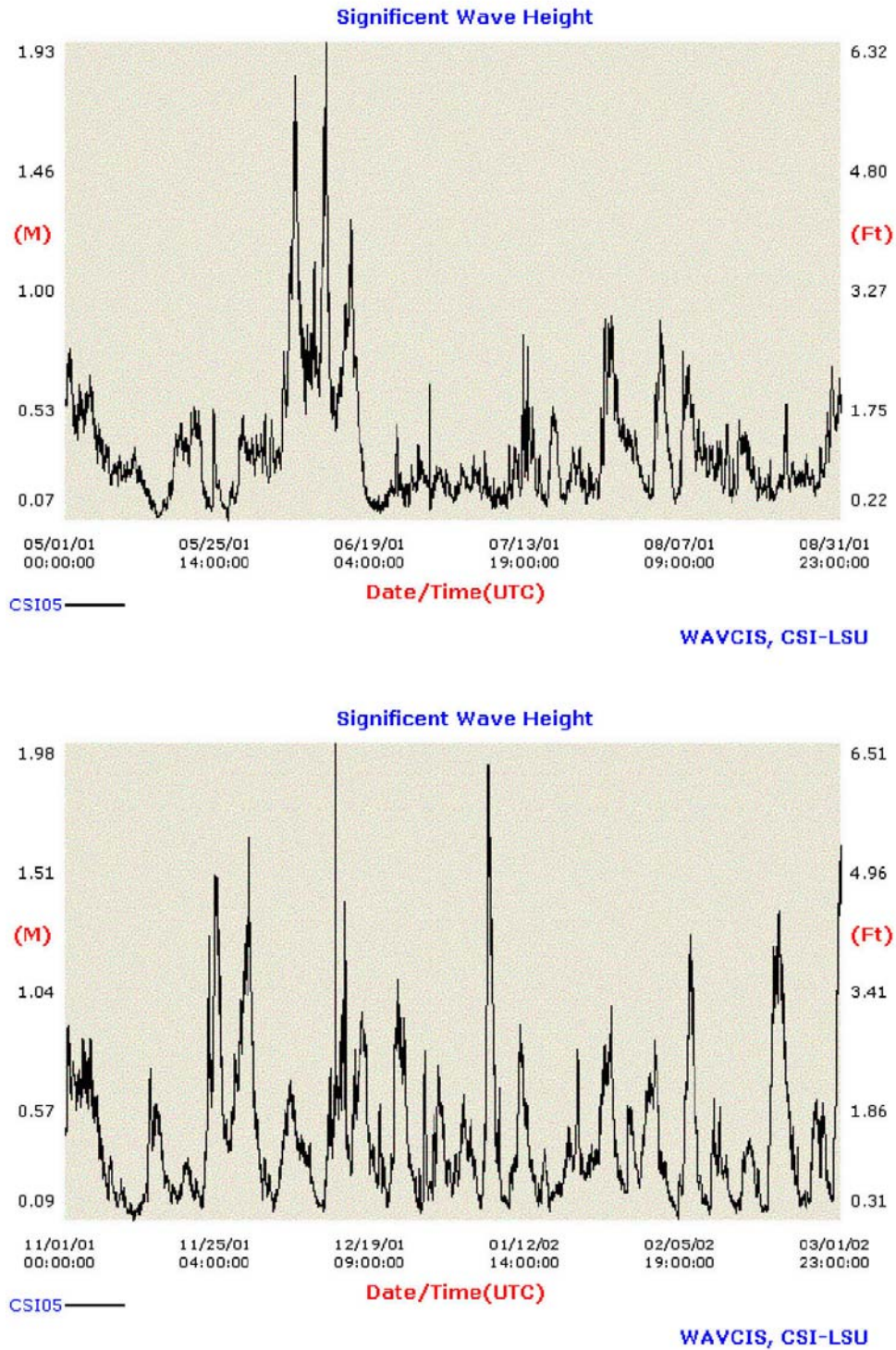


Figure D.2-3. Time series of wave height data from WAVCIS station CSI-5 during summer (upper) and winter (lower) months.

2.4 Storms

The Gulf Coast region is affected by tropical and extra-tropical storms. Tropical storms are the forerunners to hurricanes, which form when wind velocity surpasses 119 km/hr (74 mph). Hurricane season begins on June 1 and ends on November 30. During the past 100 years, 55 hurricanes and tropical storms have made landfall along the Louisiana coast with the highest incidence occurring in September (Stone et al. 1997). Extra-tropical storms form above the tropics (above the Tropic of Cancer in the Northern Hemisphere) and although they do not have the intensity of hurricanes or tropical storms, they occur more frequently. For example, in an average year, 20 to 30 cold fronts will pass through coastal Louisiana. The details of weather fronts, tropical storms, and hurricanes are presented below.

2.4.1 Cold Fronts

Cold-fronts are defined as narrow transition zones (25 – 250 km; 15 - 155 miles) between two air masses of different densities; polar air from Canada moves south meeting the warm, often moist, air mass of the southwestern and southeastern regions. Cold-fronts move west to east and commonly extend to the Gulf of Mexico. They are characterized by spatial and temporal changes in wind speed, direction, barometric pressure, temperature, and humidity (Mossa and Roberts 1990). The pressure gradient at the front controls the intensity of the system and the transfer of energy to the coast. This energy is manifested in the form of strong winds, large waves, low water levels, and abnormal rates of shoreline erosion and sediment transport and deposition.

The frontal system can be oriented oblique or parallel to the east-west trend of the coast depending upon the initial position of the air masses and how the weather system evolves (Figure D.2-4). As illustrated in Figure D.2-5, an eastward migrating low pressure system will generate an oblique frontal system. Alternatively, when a deep arctic air mass descends toward the Gulf of Mexico, a frontal system parallel to the coast develops. The type of cold front controls the duration of the storm and magnitude of the coastal processes.

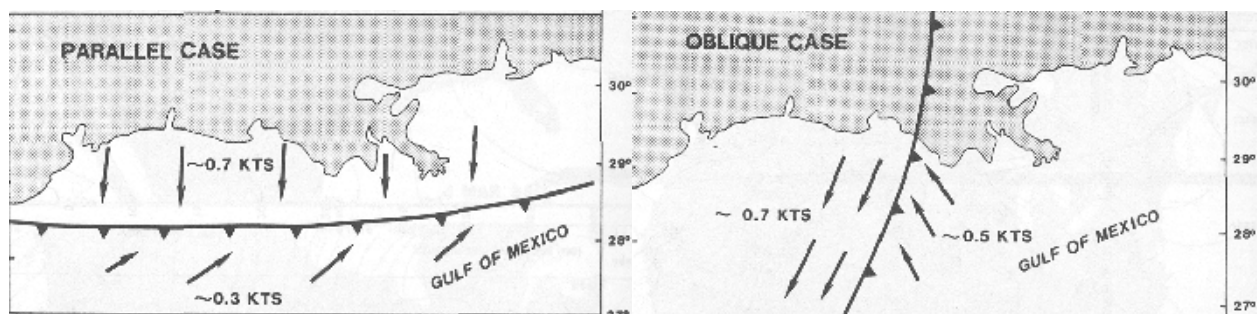
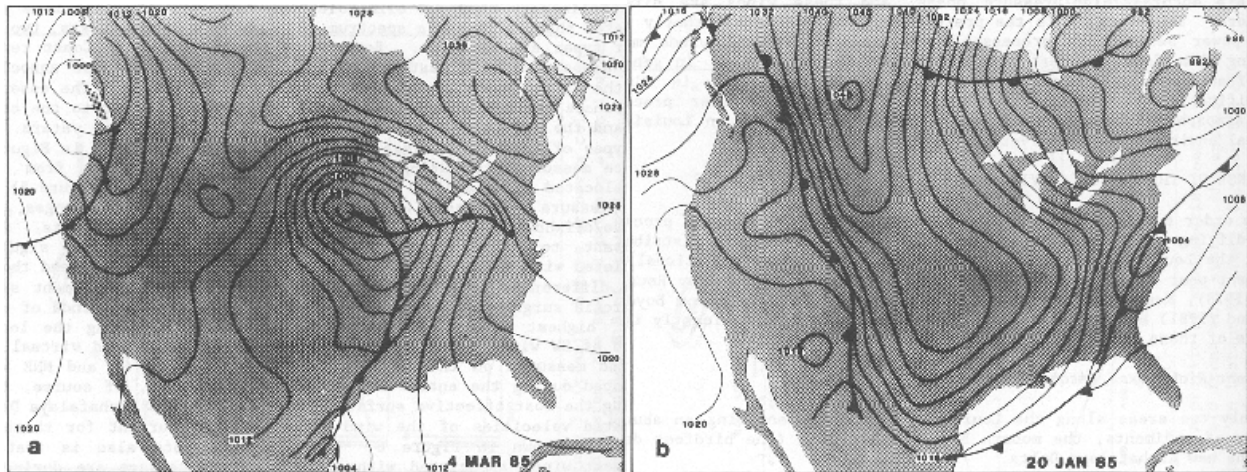


Figure D.2-4. Types of cold fronts produced by different weather systems.

In the developmental history of some coastal segments of Louisiana, cold fronts have been more important than the occasional hurricane when it comes to coastal erosion, primarily because they occur so often. Cold fronts have a typical duration of 12 to 24 hours depending on the speed of storm. Studies by local scientists characterize the cold fronts' passage as having three phases: the prefrontal, frontal passage, and the cold air outbreak, or post-frontal phase

(Roberts et al. 1987). Initially a pre-frontal phase occurs, during which, strong, warm, moist winds blow from the southerly quadrant. The resulting frontal phase is characterized by a sudden drop in air pressure, erratic winds, and short-lived, but occasionally intense, squalls. Finally, a post-frontal phase occurs, during which temperature and humidity drop, air pressure rises, and winds are strong and blow from the north. The resulting response of the coastal waters is the initial increase in tidal amplitudes, which causes waves to break higher on the beach, overwashing low barrier islands. Elevated tides increase the flow of ocean water into the bays and marsh systems behind the barrier islands. As floodwaters reside and exit the inlets with passage of the front, abrupt changes in wind direction from southerly to northerly cause large waves in the bays. This continuous process is believed to be responsible for the chronic shoreline erosion behind the barrier islands.



(a) eastward migrating cyclone type, and (b) arctic surge type cold fronts.

Figure D.2-5. Weather maps illustrating the end member types of Gulf Coast cold front passages.

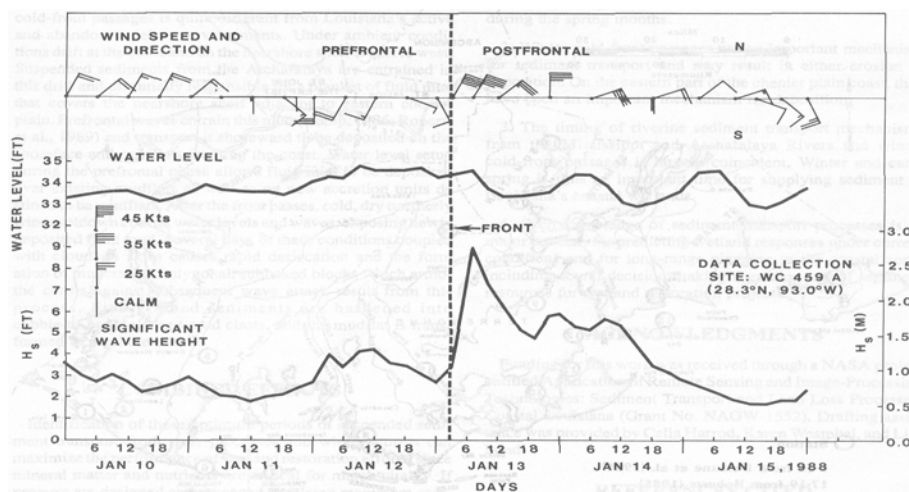


Figure D.2-6. Wind, wave, and tidal elevation data measured at a production platform off the delta coast.

Data collected in 1988 at an offshore production platform (28.3oN, 93.0oW) illustrate the effect that frontal systems can have on the coastal zone (Figure D.2-6). Prior to the cold front, the wind was from the north at 10.3 m/s (20 knots) and the significant wave height was 0.6 to 1.2 m (2 - 4 ft). As the front approached, the winds shifted to the south at 10 m/s (20 knots) but subsided as the front moved past the station. Water levels increased by 0.3 to 0.4 m (1 - 1.3 ft). Immediately following the passage of the front during a five hour period, the winds blew strongly from the north at 18 - 23 m/s (35 to 45 knots). During this time, the wave energy increased, reaching a peak significant wave height of 2.7 m (8.8 ft). Within 12 hours, the seas decreased to 1.5 m (4.9 ft), maintaining this height for another 24 hours.

2.4.2 Hurricanes and Tropical Storms

Coastal Louisiana is low-lying, with many areas of moderate population containing numerous dwellings, buildings, and other infrastructure. Destruction to natural environments, damage of property, and injury or even death to people, are all typical hurricane impacts to the coastal zone. The tracks of hurricanes through the Gulf of Mexico from 1886 to 1996 illustrate that Louisiana has been riddled with hurricanes during the past 110 years (Figure D.2-7). A graph of the spatial and temporal distribution of large magnitude storms for the 1901-1996 period indicates that hurricanes occur slightly less frequently than tropical storms, and they impact the southwest, south-central, and southeast sections of the coast more than eastern Louisiana (Figure D.2-8). The graph also demonstrates that during the past 100 years, hurricane activity peaks in September, which is consistent with hurricane activity in the Gulf of Mexico and central Atlantic Ocean (Stone et al. 1997). Similarly, 60% of tropical storms making landfall along the Louisiana shoreline take place during the months of August and September, and 80% of hurricane landfalls occur during the same period (Stone et al. 1997)..

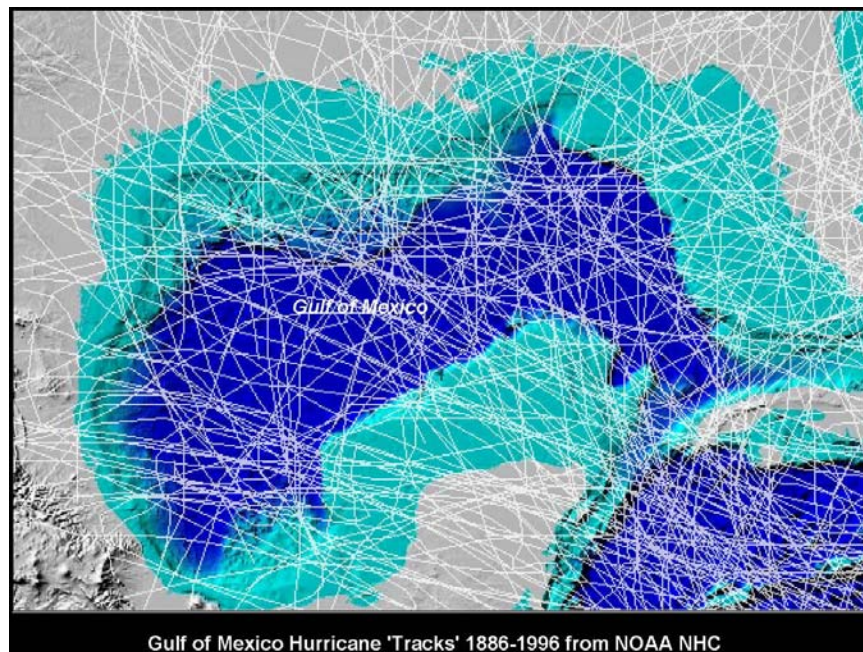


Figure D.2-7. Hurricane storm tracks in the Gulf of Mexico for the 1886-1996 period (Stone et al. 2003)

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Factors affecting the severity of a storm and the magnitude of damage to the barrier coast, headlands, and adjacent wetland areas include the strength, speed, and size of the storm. In addition, the gradient and width of the inner continental shelf are important considerations. All of these variables contribute to the size of the waves, magnitude of the storm surge, and overall impact of the hurricane. The storm surge, or storm tide as meteorologists frequently call it, is the super-elevation of the ocean water surface above the predicted tide level. It is the storm surge that allows high energy storm waves to break high against a dune ridge, across a barrier island, or over a seawall. A brief description of the factors affecting storm severity is given above (Table D.2-2).

Table D.2-2. Factors Affecting Storm Severity

Magnitude	Encompasses both hurricane size and intensity. Hurricane size governs the length of coast that is affected by the storm as well as the duration of high velocity winds and high energy waves. The greater the intensity of the hurricane, the stronger the wind velocities and the larger the wave heights and storm surge
Speed of Storm	Determines amount of time that storm winds can transfer its energy to the water surface waves and pile water onshore. Generally, slower-moving storms produce higher waves and larger storm surges than faster moving storms of equal magnitude
Path of Storm	Determines the landfall of a hurricane and areas along the coast of greatest storm impact. In the northern hemisphere when a hurricane moves onshore, areas to the right of its landfall will experience the strongest winds and greatest storm damage
Coastline Configuration	An important factor in large deeply embayed coastlines. In this setting certain hurricane tracks can significantly amplify storm surge levels as water is forced into the funnel-shaped embayment

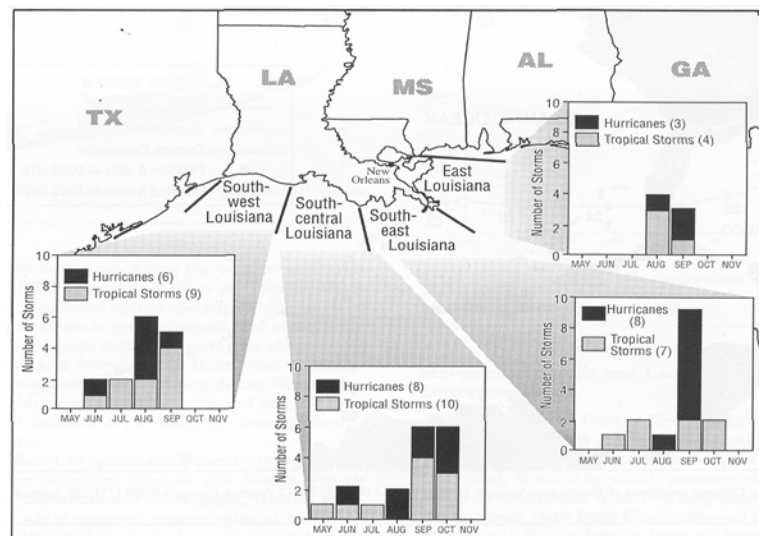


Figure D.2-8. Graph showing distribution of hurricanes and tropical storms along the Louisiana coast from 1901 to 1996 (from Stone et al. 1997).

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Hurricanes are one of earth's greatest demonstrations of physical energy. A satellite image on 26 September, 1998 shows Hurricane Georges covering much of the Gulf of Mexico and the coastal states of Louisiana, Mississippi, Alabama, and Florida (Figure D.2-9). The swirls of clouds radiating from the storm's center depict the internal structure of a hurricane. The air mass of the low-pressure system flows in a counter-clockwise direction at the base of the storm. Toward its center, the air mass spirals upward and eventually flows outward at the top of storm. The counter-clockwise circulation of these storms explains why when a hurricane approaches the shores of Louisiana it is the coastal region along the eastern half of its landfall that experiences the strongest winds, highest storm surge, and greatest damage.

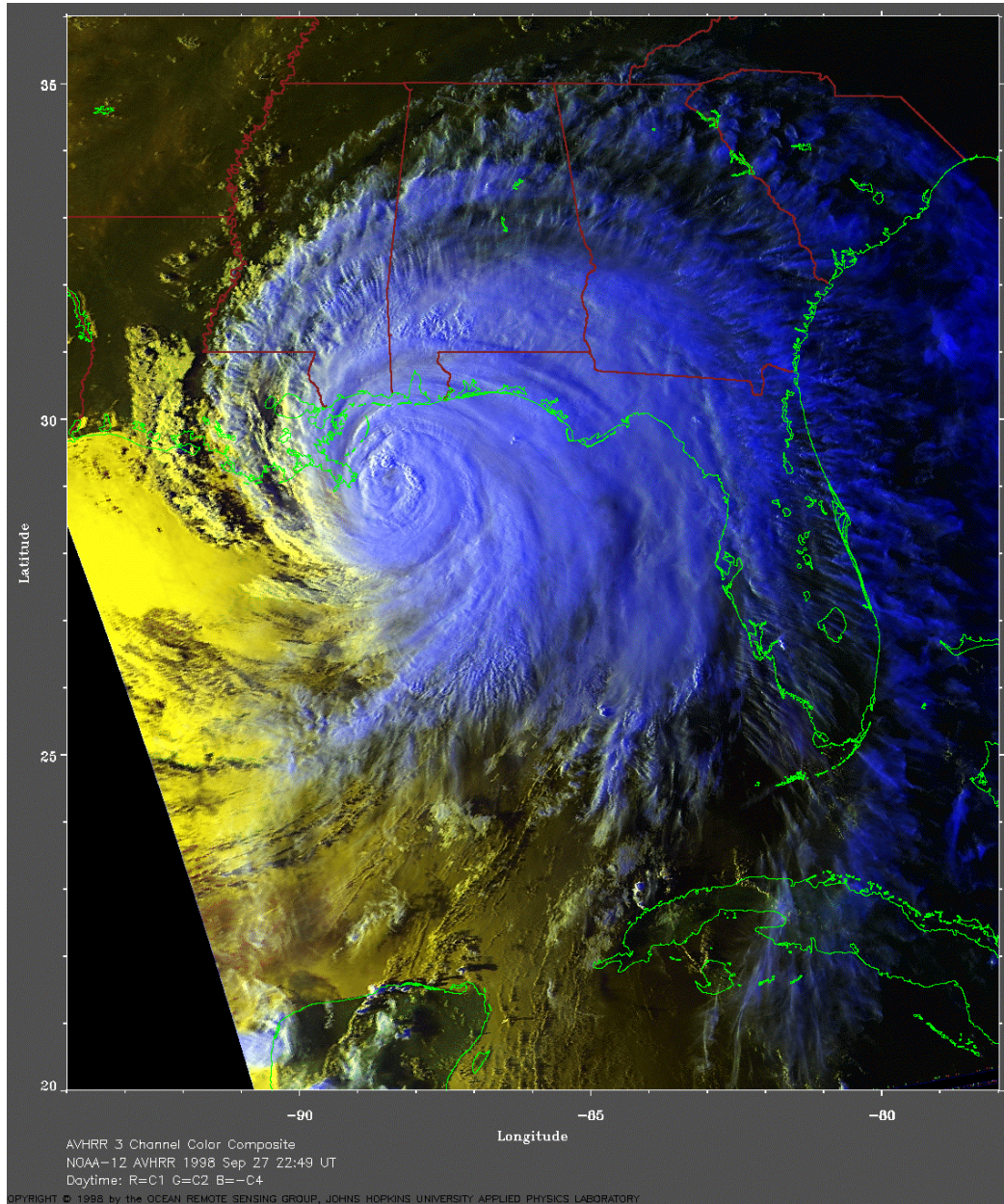


Figure D.2-9. Satellite view of hurricane Georges before landfall on September 27, 1998 off coastal Louisiana.

The strength of a hurricane is formally classified using the Saffir-Simpson Scale and is based on maximum wind velocity, barometric pressure, storm surge level, and expected damage (Table D.2-3). The scale has five categories, each representing a substantial increase in hurricane intensity. Pathways of different category hurricanes impacting Louisiana are depicted in Figure D.2-10. Note the good correlation between the category of a hurricane and its storm surge as measured at the Gulfport, Mississippi tide gage station (Figure D.2-11). Hurricane Camille was a category 5 storm that had wind velocities in excess of 322 km/hr (200 mph) producing a surge of 6 m (20 ft). In contrast, Hurricane Georges was a category 2 storm with wind velocities less than 160 km/hr (100 mph) and it generated only a 2.2 m (7 ft) surge.

Table D.2-3. Storm Rating based on the Saffir-Simpson Scale

Category 5	< 250 km/hr (155 mph)
Category 4	211 - 250 km/hr (131 - 155 mph)
Category 3	180 - 210 km/hr (111 - 130 mph)
Category 2	155 - 179 km/hr (96 - 110 mph)
Category 1	120 - 154 km/hr (74 - 95 mph)
Tropical Storm	63 - 119 km/hr (39 - 73 mph)
Tropical Depression	< 63 km/hr (39 mph)

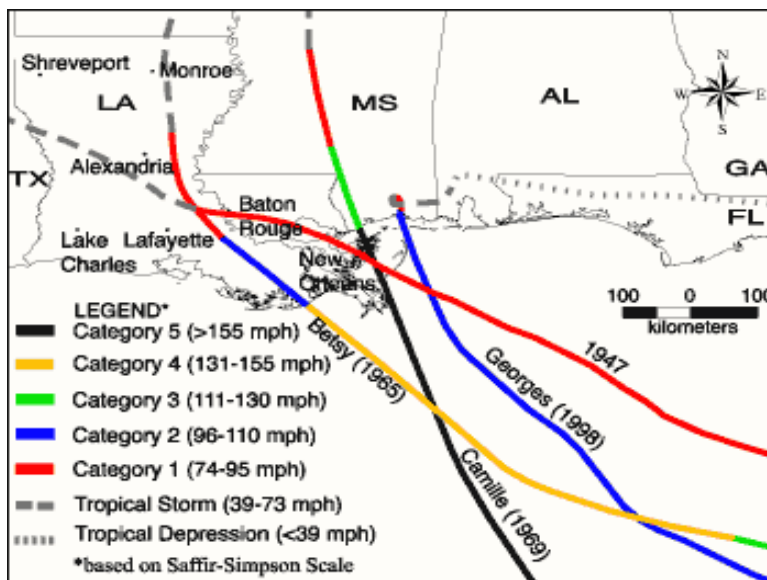


Figure D.2-10. Tracks and Category Status of Hurricane 1947, Betsy, Camille, and Georges.

Recently, Louisiana experienced the impact of two major storms that caused widespread flooding, shore erosion, and barrier overwash and breaching. Tropical Storm Isidore made landfall along the western Louisiana coast at Marsh Island on the 26 September, 2002 and seven days later on 3 October, 2002, Hurricane Lili made landfall at Bayou Lafourche in central Louisiana (Figure D.2-12). Ocean parameters for these storms were recorded at the WAVCIS station (CSI-5) (Figure D.2-13). According to CSI Station 5, Isidore produced a storm surge of 0.6 m (2 ft), and Hurricane Lili produced a 1.2 m (4-foot) storm surge along the coast (Figure D.2-13). Likewise, the significant wave height for Isidore was 2.3 m (7.5 ft) compared to 2.8 m (9.2 ft) for Lili (Figure D.2-14). Hurricane Lili produced significant erosion and overwash along the Isles Dernieres and Timbalier barrier islands (Figure D.2-15). The severity of overwash was due to the fact that Tropical Storm Isidore had already eroded much of the barrier shoreline and removed the protective beach. The temporal proximity of the two events resulted in much greater change to the coast than would have been predicted if Hurricane Lili had been the only storm of the season to make landfall in Louisiana.

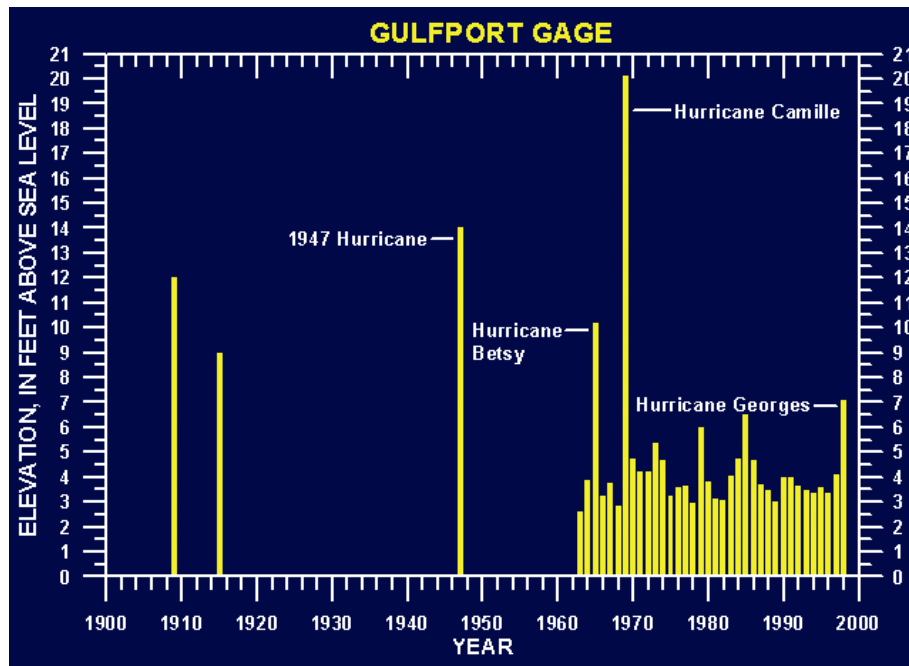


Figure D.2-11. Storm Surge produced by hurricanes 1947, Betsy, Camille, and Georges, at the Gulfport gage.

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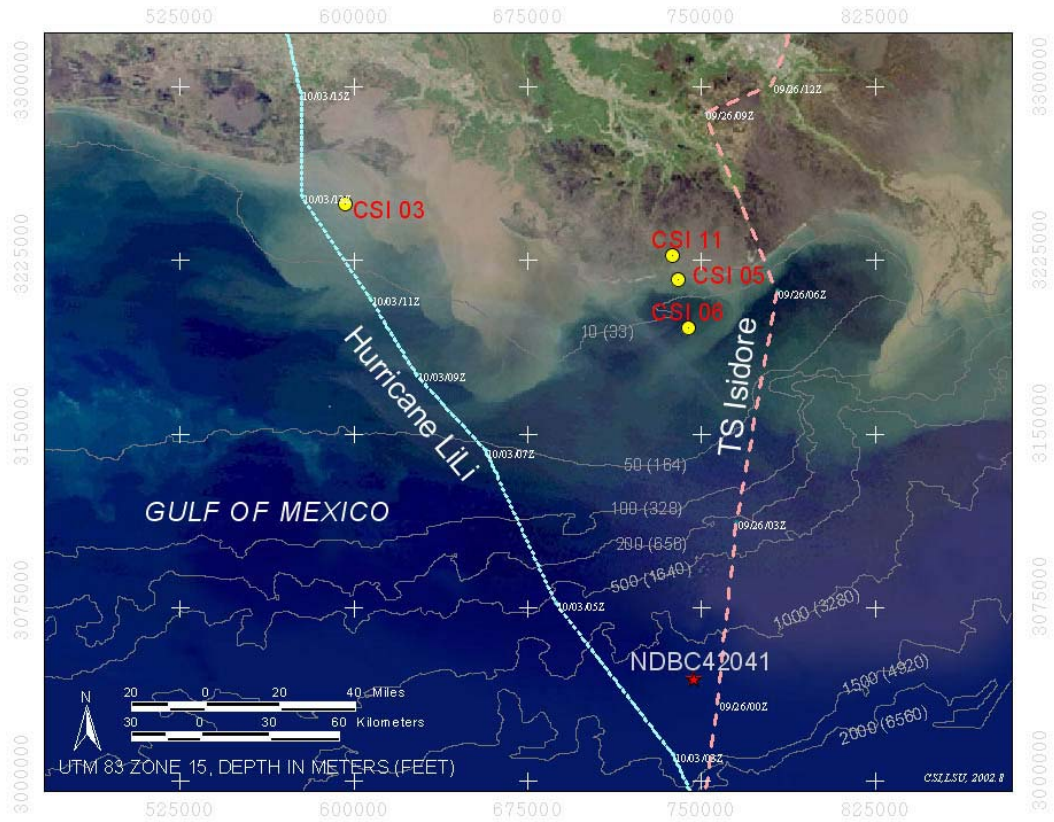


Figure D.2-12. Tracks for Tropical Storm Isidore and Hurricane Lili (from Stone and Sheremet 2003).

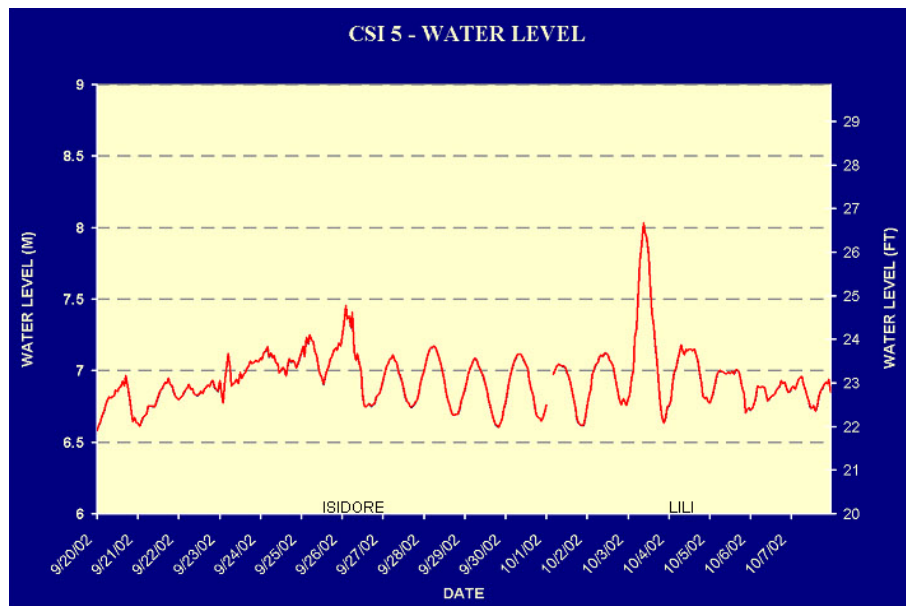


Figure D.2-13. Water levels as a result of Tropical Storm Isidore and Hurricane Lili (from Stone and Sheremet 2003).

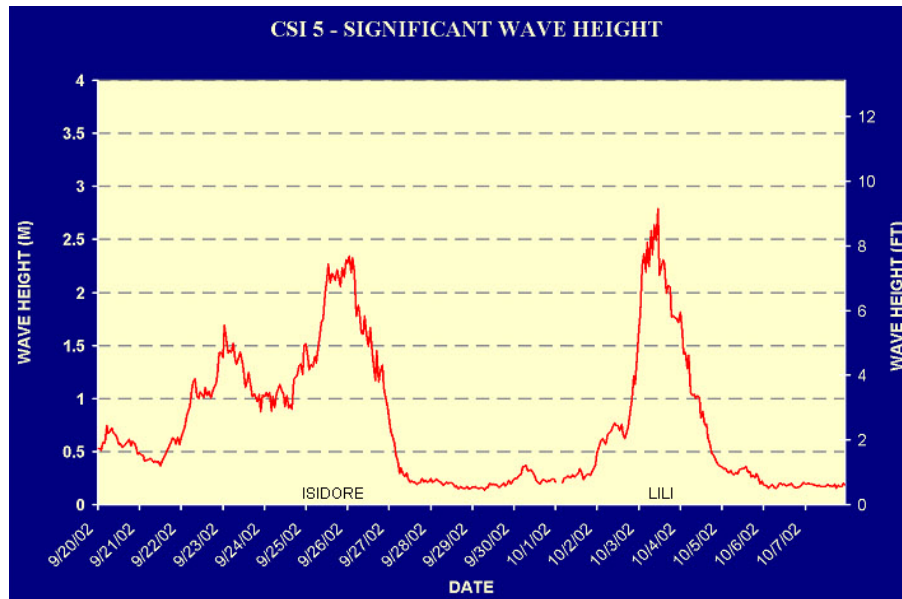


Figure D.2-14. Significant Wave Heights as a result of Tropical Storm Isidore and Hurricane Lili (from Stone and Sheremet 2003).

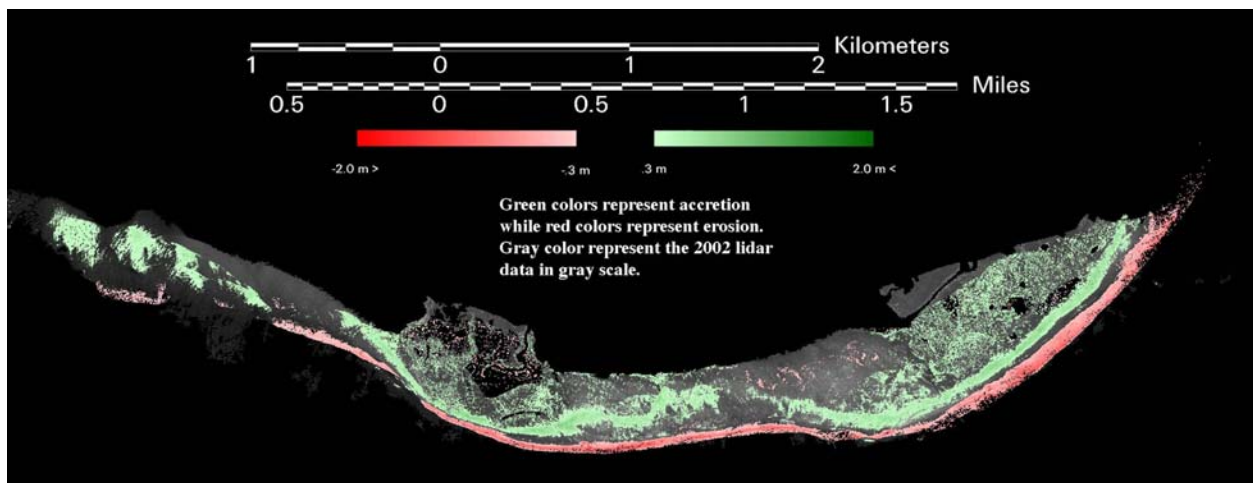


Figure D.2-15. Lidar images post-Hurricane Lili showing net accretion and erosion.

2.5 Tides

Tides along the Gulf Coast states are a manifestation of the moon and sun's gravitational attraction of earth's hydrosphere, which in this case is the waterbody encompassing the Gulf of Mexico. The rise and fall of the tides is one of the major rhythms of the earth. People familiar with the shore are aware that tides control the elevation and times of high and low tide as well as the flow of water into and out of bays, harbors, and tidal inlets. Tides interacting with freshwater discharge from rivers and produce salinity gradients in estuaries and wetland areas. Every region along the coast has its own distinct tidal signature that is a product of: (1) the tidal wave rotating in a counter-clockwise pattern in the Gulf of Mexico, and (2) perturbations of that wave as it shoals across the continental shelf and propagates into embayment and other coastal regions.

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Tides along the Louisiana coast change systematically, with an overall decrease in tidal range from the western Chenier Plain eastward toward Mississippi Sound (Figure D.2-16). At Calcasieu Pass in western Louisiana, the tides are mixed and have a strong diurnal component. As illustrated in Figure D.2-16, the first part of the time series records semi-diurnal tides that gradually become more and more unequal. By Day 23, the tides are almost diurnal, and the second low tide is really just a short stillstand during an overall period of falling tide elevation. At the beginning of a new fortnightly cycle, the tide curve exhibits a more semi-diurnal signature. Daily tidal elevations vary in range from a low of 60 cm (2 ft) to a high of 0.97 m (3.2 ft).

In the delta region including Raccoon Pass (west), Grand Isle (central), and the northern Chandeleur Islands (east) tides are strongly diurnal. At Raccoon Pass, the tidal range varies from a low of 15 cm (0.5 ft) during equatorial tidal conditions (Day 27, Fig. D.2-16) to a high of 97 cm (3.2 ft) during tropic tides (Day 17, Fig. D.2-16). In the central and eastern portion of the delta the tidal wave is dampened, resulting in smaller equatorial tidal ranges (TR = 12 cm (0.4 ft); Day 27, Fig. 15) and smaller tropic tidal ranges (TR = 64 cm (2.1 ft); Day 17, Fig. D.2-16) as compared to the Calcasieu tides.

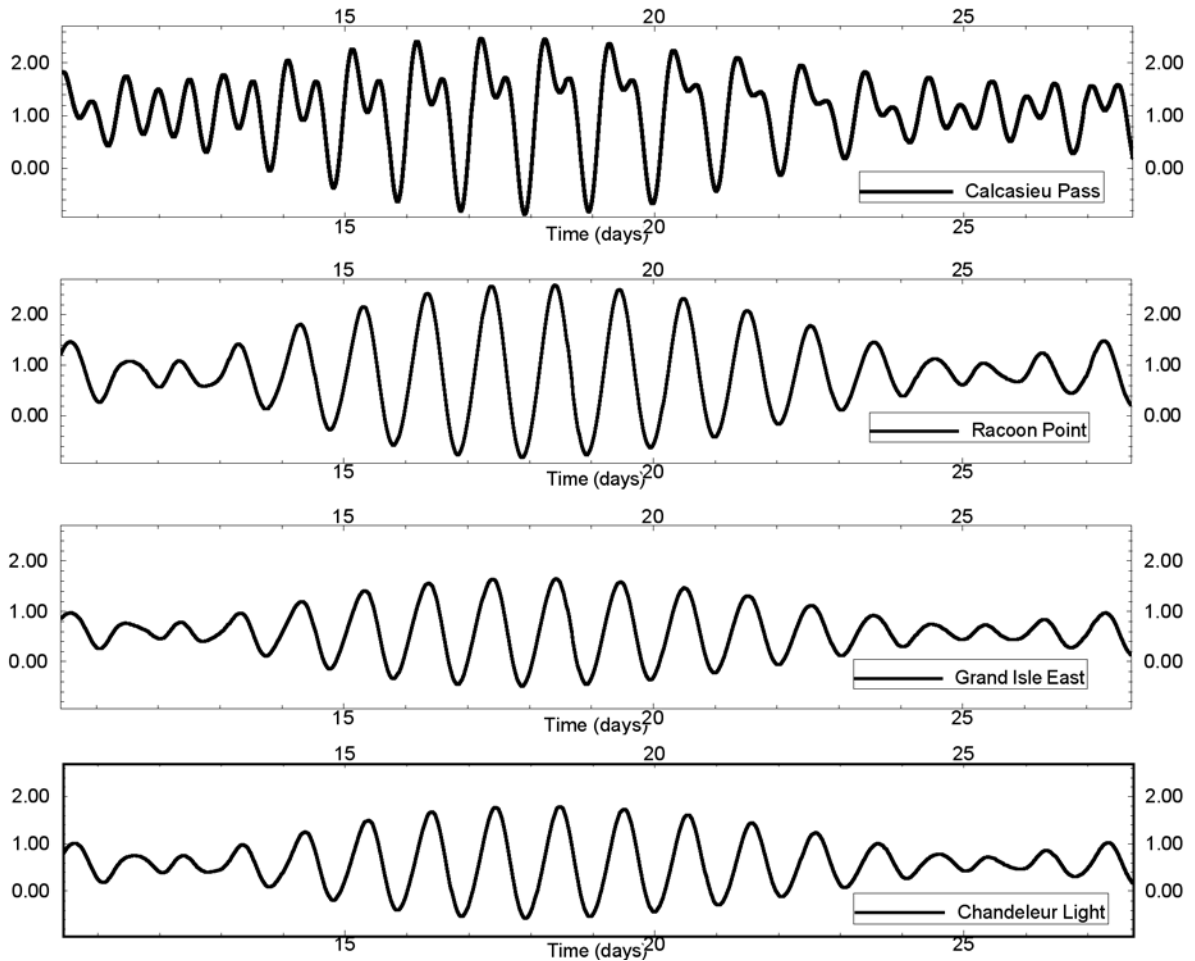


Figure D.2-16. Typical tidal signature along coastal Louisiana from west (top) to east (bottom).

2.6 Sea Level Rise and Subsidence

The trend of rising sea-level along Louisiana's coast is attributed to several mechanisms, both natural and anthropogenic. Essentially, the land surface is sinking (subsiding), and the ocean level is rising. Melting glacial ice is adding more water to the ocean basins, and the ocean waters are warming and expanding. Together these processes produce a sea-level signature that is specific to a particular section of coast and termed "relative sea-level rise." By recording the elevation of sea level at a given location over time, we are able to generate a relative sea-level curve for that site. Subsidence of the Mississippi River Delta is related to natural processes, such as sediment compaction, faulting, and isostatic adjustment to regional crustal loading. Anthropogenic related subsidence is due to the withdrawal of fluids (gas, oil, and water) (Kolb and Van Lopik 1958; Jurkowski et al. 1984; Penland et al. 1989; Morton and Purcel 2001). Recent work by the USGS suggests that removal of fluids has the cumulative effect of decreasing pore pressure, which can initiate formerly active faults. The relative role of the natural and anthropogenic factors on net regional subsidence has not been well established, despite their potentially significant influence on coastal erosion patterns and sedimentation processes.

Historical sea-level curves exist for Eugene Island along the west delta coast and for Grand Isle along the central delta coast (Figures D.2-17 and D.2-18). The curve for Grand Isle extends back to the late 1940s and continues to the present time, whereas the Eugene Island curve covers only the period between the early 1940s to the mid-1970s. The Grand Isle curve covers a longer period, extends to the present, and is considered most reliable for determining historical trends and forecasting potential sea-level rise scenarios. Data for Grand Isle indicate that relative sea level is rising at a rate of 1.03 cm per year (0.4 inches/year; Penland and Ramsey 1990). This is the highest rate along the contiguous United States as illustrated in Figure D.2-19 and helps to explain coastal evolution in Louisiana.. Subsidence and rising sea level are also largely responsible for shoreline erosion and the transgressive nature of most of the barrier islands in Louisiana. The historical decrease in size and increasing segmentation of Louisiana's barrier systems as well, as the loss of wetlands, has greatly increased the vulnerability of the mainland areas during major hurricanes.

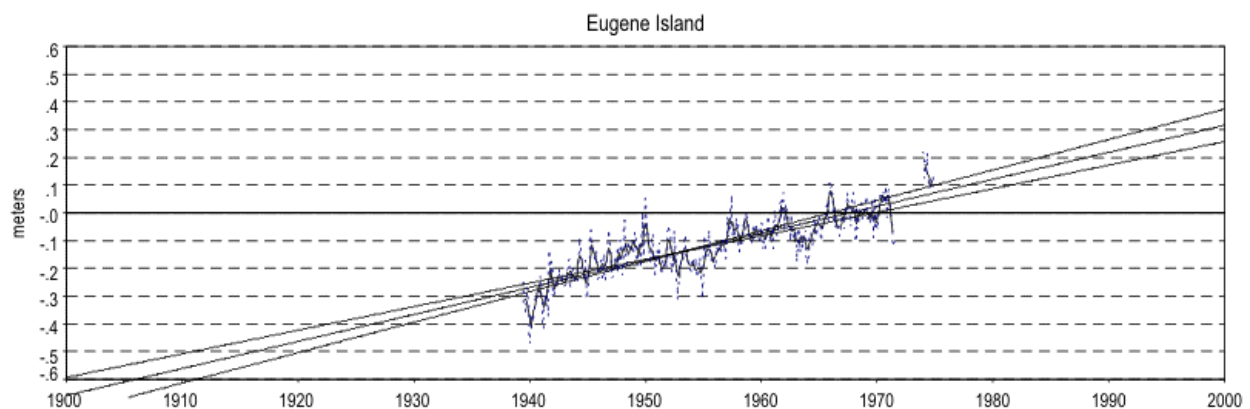


Figure D.2-17. Tide Gauge at Eugene Island.

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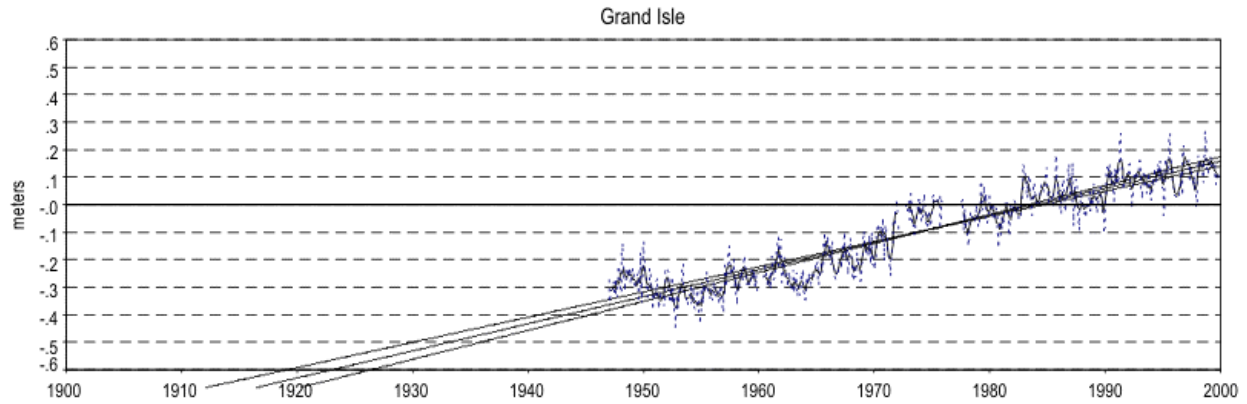


Figure D.2-18. Tide Gauge at Grande Isle.

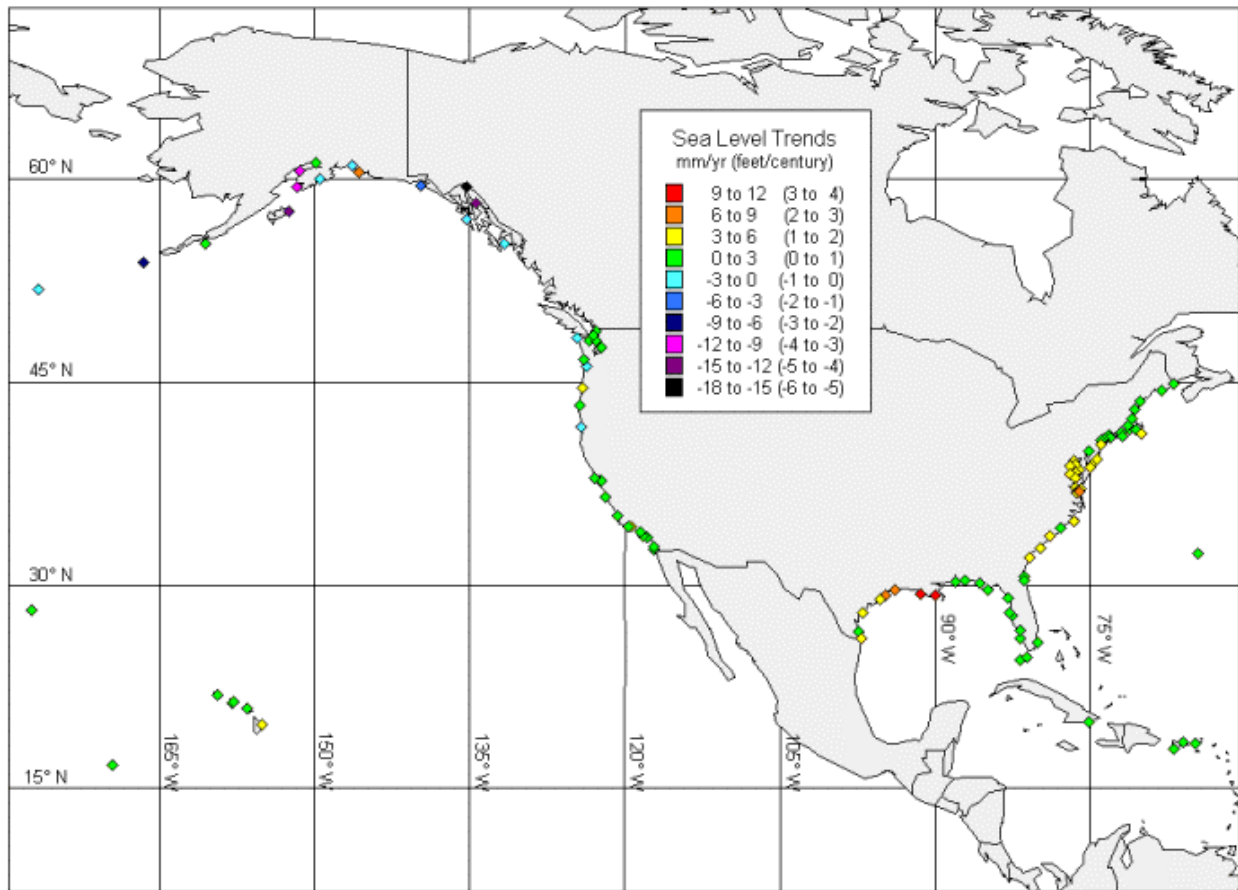


Figure D.2-19. Map of subsidence rates in North America.

2.7 Tidal Inlets and Tidal Prism Dynamics

Barrier island development along the Louisiana coast is a product of river avulsion and the subsequent reworking of distributary headlands (Penland et al. 1988). The size and number of

tidal inlets along the barrier coast are controlled, in part, by the volume of water (tidal prism) moving into and out of backbarrier bays. The historical evolution of these tidal inlets is a product of changes in extent and configuration of the backbarrier bays. Detailed geomorphic and bathymetric changes of the lower delta plain between the 1880s and the 1990s, including the barrier and tidal inlet systems, are contained in a two atlas set published by the U.S. Geological Survey (Williams et al. 1994; List et al. 1994). Levin (1993) also discusses historical development of the inlets. Generally, tidal exchange between backbarrier bays and the Gulf of Mexico has increased along the delta plain since at least the 1880s due to widespread conversion of wetlands and salt marsh to open water areas. For example, in the mid-1800s the Isles Dernieres were backed by Lake Pelto. At that time, the lake was surrounded by a near uninterrupted expanse of marshland. During the next hundred years, land subsidence, wave erosion of the marsh shoreline, and dredging activity transformed the lake into a large continuous sound having an open connection to Caillou Bay to the west and Terrebonne Bay to the east. The historical changes that have occurred to the Isles Dernieres are symptomatic of the wetland loss and barrier evolution along the entire delta plain coast.

Overall, extensive engineering along the lower Mississippi River and confinement of its discharge have dramatically decreased sediment influx to much of the lower delta, reducing the ability of the marsh to keep pace with rising sea level. Thus, the hypsometry of the backbarrier has evolved toward greater subtidal environments and less intertidal and supratidal areas. This trend has strongly affected tidal inlet geometry and sediment dispersal along the barrier complexes.

Tidal prism dynamics and the pattern of tidal exchange dictate the occurrence and geometry of tidal inlets along the various barrier chains. The northern and southern Chandeleur Islands front Chandeleur and Breton Sounds, respectively. Both of these broad bays have an open connection to the Gulf of Mexico, and thus most of the tidal waters pass around the barrier island chain rather than through tidal inlets. Many of the inlets along the Chandeleur Islands are formed during storms and are ephemeral in nature. Generally, inlets of this section of coast are small in size due to small tidal prisms. Tidal inlets along the Timbalier Islands and Isles Dernieres have highly variable geometries due to the segmented nature of these barrier systems. Much of the tidal exchange between the backbarriers of Caillou Bay, Terrebonne Bay and Timbalier Bay and that of the Gulf of Mexico occurs through broad shallow channels where the transgressive barriers have undergone extensive erosion. However, there are several relatively deep passes (6 to 10 m; 20 - 33 ft) that are maintained by strong tidal currents (~ 1.0 m/sec; 3.28 ft/s).

The barrier chain that has formed between the Caminada-Moreau headland region and the Plaquemines delta lobe is somewhat different from the other barrier chains because it is more robust, less segmented, and tidal flow into and out of its backbarrier occurs entirely through well-defined tidal inlets. During the past half-century, rapid relative sea-level rise (1.03 cm/yr; Penland and Ramsey, 1990) and other erosional processes within Barataria Bay have led to substantial wetland loss, converting more than 775 km² of wetlands to open water (Figure D.2-20; Barras et al. 1994). As the open water area increased, so has the bay tidal prism and tidal exchange. Between 1880 and 1990, the enlarging tidal prism produced a 44% increase in the combined cross-sectional areas of the major tidal inlets of Barataria Bay (Figure D.2-21). The increase in size of the tidal inlets was at the expense of the adjacent barrier islands. During the same period of time there was concomitant progradation of the ebb-tidal deltas. For example,

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since the 1880s the ebb delta at Barataria Pass built seaward more than 2.0 km (1.25 miles). Coastal erosion and increasing bay tidal prism also facilitated formation of new tidal inlets, including Pass Abel.

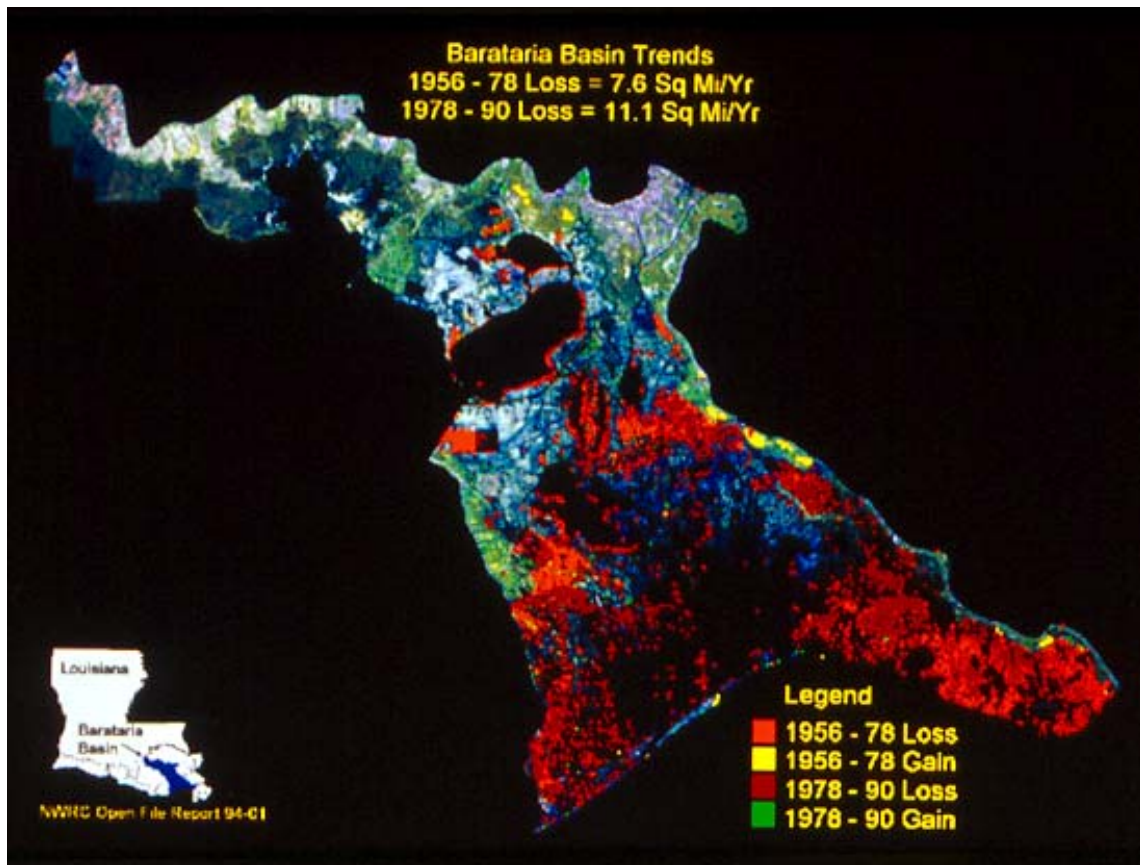


Figure D.2-20. Map of Barataria Basin land loss/gain.

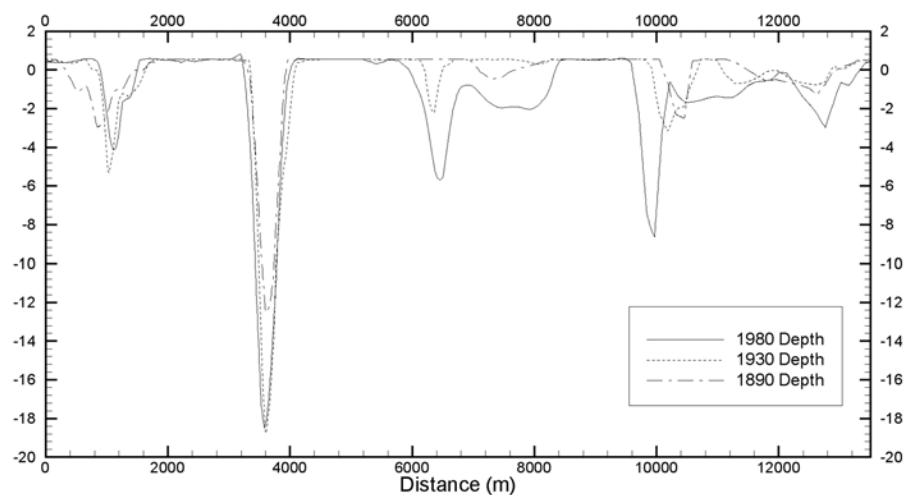
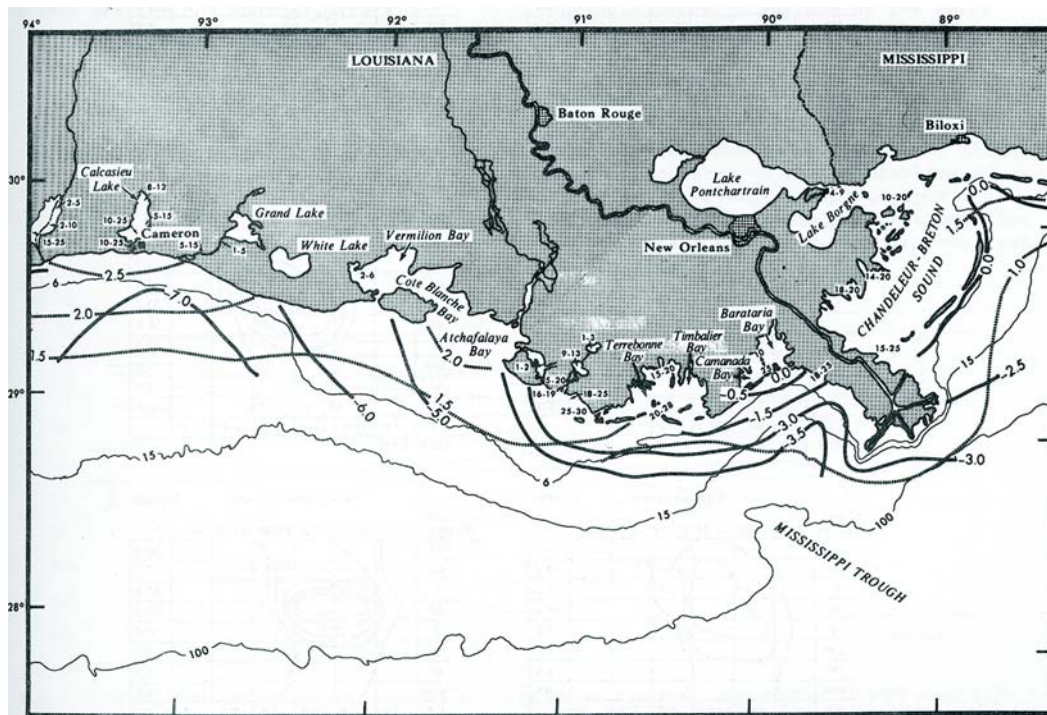


Figure D.2-21. West to east cross sections at Caminada Pass, Barataria Pass, Pass Abel, Quatre Bayou, showing inlet changes since 1890..

2.8 Estuarine Circulation

Circulation of coastal waters depends on driving forces such as tides, wind, and atmospheric pressure. Along the complex Louisiana coast, circulation mechanisms go beyond these driving forces to include high rainfall; the large volume of fresh water introduced by the Mississippi and Atchafalaya Rivers; currents induced by density differences and mixing processes of these two masses of water; local shoreline and bathymetric features such as the Mississippi River mouth, barrier islands, marshes, inlets, bays, and so forth.

Tidal currents in Louisiana are relatively small, due to the small tidal amplitude. In the absence of wind, density effects and barometric pressure gradients, these currents reach magnitudes of approximately 10 – 15 cm/s (0.3 - 0.5 ft/s). Estimates and observations suggest that tidal currents are stronger at the surface water column and are decreasing with increasing depth. This occurrence is primarily due to the encounter of denser and heavier salty gulf waters in deeper regions, which are less likely to respond to small tide variations. Although small in magnitude in open coastal waters, tidal currents can reach speeds of approximately 50 cm/s (1.7 ft/s) at estuary and barrier island inlets, depending on the inlet dimensions. The amount of circulation attributed to rising and falling tides or tidal induced circulation is measured as a function of the spatial and temporal variability of tides along the Louisiana coast. This variability is shown in Figure D.2-22 (Murray 1976). There is a seven hour lag before high water from the east coastal zone reaches the west coastal zone, with typical tidal ranges between 30 - 60 cm (1 - 2 ft) depending on the time of month and year.



Co-phase lines of the tide in hours before high water at Barataria Pass are shown as heavy solid lines. Co-range lines of the tide in feet are shown as dashed lines. Thin solid lines are water depths in meters. Small numbers indicate observed salinity ranges (ppt) in bays and estuaries (after Murray 1976).

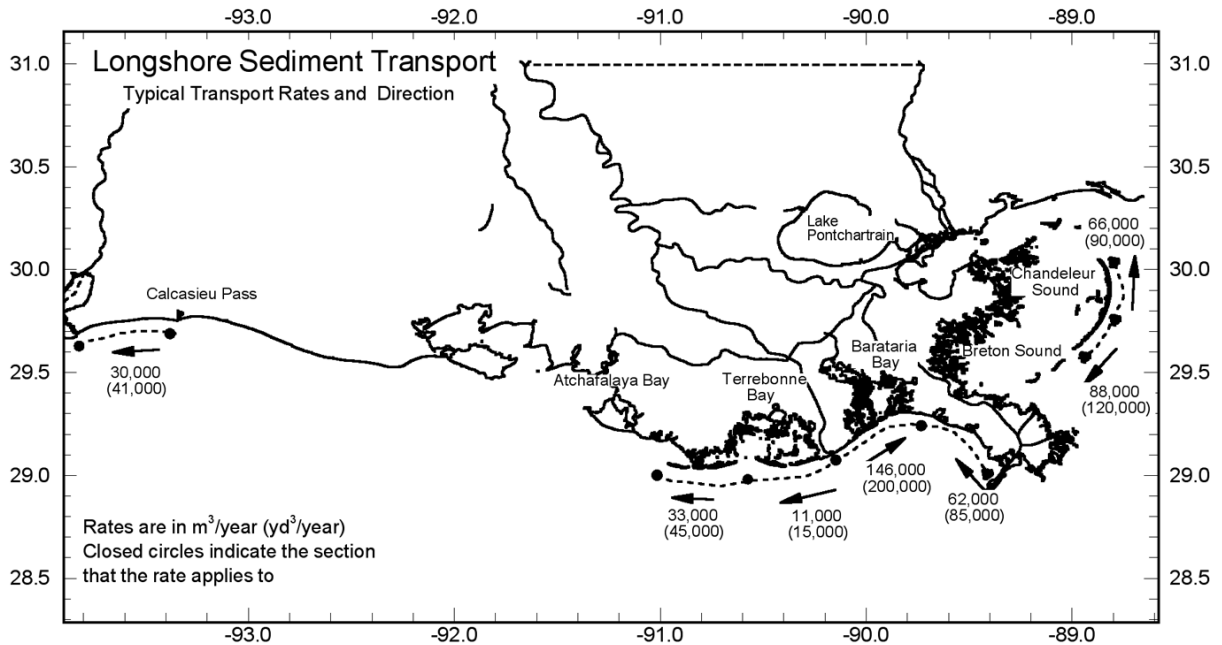
Figure D.2-22. Location map of Louisiana coastal waters.

Perhaps more critical than tides, in terms of circulation and mixing, are wind and barometric pressure. Wind can induce circulation in the form of set-up and set-down, seiche, and wind-waves. Similarly, the presence of front-like weather during the winter and storms during hurricane season enhances these processes by producing dynamic wind conditions. The speed and direction of these winds shift abruptly, creating strong gusts. Changing wind speed and direction cause the generation and transformation of waves along the Louisiana coast. Wind and barometric pressure induced circulation is critical and dominant in back bays, enclosed bays, and lakes, marshes, and sub-tidal areas. These processes are characterized by extreme water level fluctuations, and are responsible for most of the erosion taking place along the Louisiana coast.

Another important process in Louisiana is the freshwater exchange and mixing attributed to the Mississippi and Atchafalaya Rivers, and the hundreds of streams and bayous along the coast. The two rivers combined account for a mean flow of approximately 20,000 m³/s (700,000 cfs), with seasonal variations of up to 8,000 m³/s (280,000 cfs). The low density freshwater meets and mixes with the higher density Gulf waters. In the process, the density difference between the two water masses causes density currents. Typically the surface water column is fresh or brackish depending on the strength of the discharge, and the bottom water column is salty. The speed of the density currents varies in magnitude and is a function of the density difference, discharge velocity, and available head difference. Although relatively small for the most part, these currents can propagate several miles upstream or downstream given the presence of favorable conditions. Periodic intrusion of saltwater can be detrimental to critical habitat in the bays and marshes. Episodic exposure to highly saline water leads to marsh deterioration and systematic land erosion. Salt water intrusion in the Mississippi River has been observed to travel more than 80 km (~50 miles) upstream during low flow. Similarly, during periods of low rainfall and hence low fresh water discharge, salt water wedges slowly propagate onshore for several miles through inlets and bays.

2.9 Longshore Sediment Transport

Longshore sediment transport is the movement of sediment parallel to the shore. This process is a result of breaking and shoaling waves suspending sand from the bottom and the displacement of the sediment downdrift by the longshore current. The magnitude of the longshore current increases with increasing wave height and breaker angle. In addition to these wave parameters, the rate of transport is a function of beach or barrier orientation, offshore shelf slope, and local depth. In coastal Louisiana, direct measurements of longshore transport are limited. The rates of transport are typically based on historical studies of shoreline erosional and depositional trends, sedimentation patterns in the vicinity of coastal structures, and on numerical wave modeling. An overview of the general trends of sand movement along the Louisiana coast is shown in Figure D.2-23 and Table D.2-4. Sediment transport trends along the coast are discussed in terms of shoreline segments based on barrier island arcs and other geomorphic features.



Rates are in cubic meters per year (m^3/yr) with cubic yards per year (yd^3/yr) in parenthesis. Arrows indicate net dominant transport direction and closed circles indicate the section that the given rate applies to.

Figure D.2-23. Longshore sediment transport estimates in coastal Louisiana

Table D.2-4. Longshore Sediment Transport Rates for Coastal Louisiana

<u>Geographic Location</u>	<u>Coverage</u>		<u>Rate</u>	
	<u>From:</u>	<u>To:</u>	<u>(yd^3/yr)</u>	<u>Direction*</u>
Holly Beach Area	Calcasieu Pass	Sabine Pass	41,000	Westward
Isle Dernieres Chain	East Island	Racoon Island	45,000	Westward
Timbalier Chain	Racoon Pass	Cat Island Pass	15,000	Westward
Caminada Moreau Headland	Belle Pass	Grand Pass	200,000	Eastward
Sandy Point	South Pass	Grand Pass	85,000	Westward
Chandeleur Islands	Southwest flank	Island Center	120,000	Westward
Chandeleur Islands	Island Center	Northeast flank	90,000	Eastward

2.9.1 West Louisiana - Holly Beach (Calcasieu Sabine)

The net longshore sediment transport from Calcasieu Pass to Sabine Pass is predominantly westward, with the exception of localized reversals at both passes (Underwood et al. 1999). The transport reversal is attributed to wave refraction caused by offshore shoals (Sabine Bank) and the dredge mounds on the flanks of the channels. Transport rates generally increase west of Calcasieu Pass toward Constance Beach where a maximum rate of 30,000 m³/yr (~41,000 yd³/yr) occurs (Figure. D.2-24. Moving west from Constance Beach to Sabine Pass, the sand transport rates decrease.

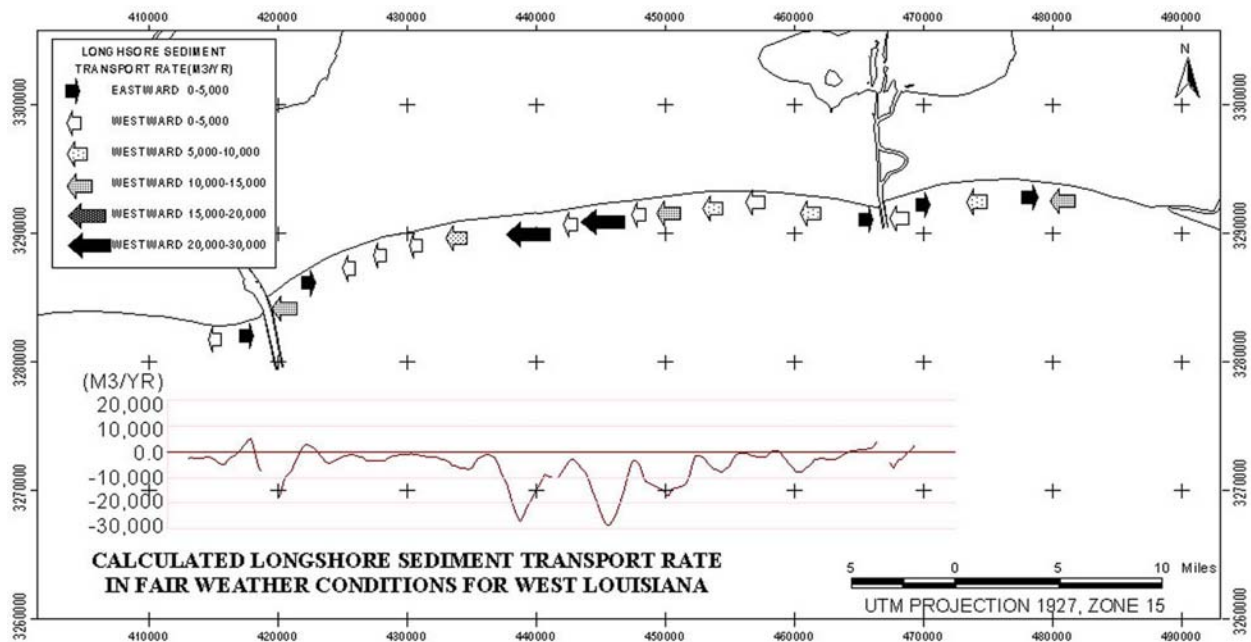


Figure D.2-24. Calculated longshore transport rates in west Louisiana under fair weather conditions (Underwood et al. 1999).

2.9.2 Isles Dernieres

Sediment transport along the Isles Dernieres is complex given its fragmented nature (Stone and Zhang 2001a). Overall, sediment moves in a westerly direction along the Isles Dernieres island chain, although local bi-directional transport occurs on Trinity and Whiskey Islands. Sediment transport is largely inhibited across Whiskey Pass due to the dominance of flood currents and landward transport through the inlet. Waves propagating through the pass break along the marsh shoreline in Lake Pelto (Stone and Zhang 2001a). This process indicates that sand is transported predominantly onshore through the pass, thereby minimizing sediment bypassing to downdrift Whiskey Island.

2.9.3 Timbalier Islands

Net sediment movement along the Timbalier Islands is to the west, and the rate increases from east to west (Stone and Zhang 2001b). Sub-scale transport trends are evident on both islands. East Timbalier Island is dominated by westward transport, with a net increase in rate to the west. However, the sand transport system along the island has been greatly diminished due to the extent of coastal structures in the area. The potential for transferring sand from the Caminada Moreau headland to East Timbalier Island is minimal, given the large width of Raccoon Pass and the net landward transport of sand to its flood tidal delta. Kulp et al. (2003) have documented extensive growth of this flood tidal delta during the past ten years. This suggests that little sand bypasses the inlet. Rather, the sand is worked onshore into Timbalier Bay. Similarly, net transport is westward along Timbalier Island with a net increase in rate along the eastern flank of the barrier island. Conversely, the rate decreases to the western end of the island. This pattern suggests that sand eroded from the eastern flank is transported to the west where it is deposited along the west flank of the barrier and in Cat Island Pass. Bypassing of sand across Little Pass Timbalier is minimal. Waves propagate through this inlet prior to breaking inside Timbalier Bay. In addition, dense armoring along East Timbalier Island diminishes the longshore transport of sediment out of this system to the west.

2.9.4 Chandeleur Islands

Sediment transport along the Chandeleur Islands is known from wave modeling studies by Ellis (1998) and Stone and Ellis (in press) (Figure D.2-25). There is a sediment drift nodal point located in the south-central portion of the island. North of this point, sand moves northward for 25.5 km (16 miles) along the barrier system. South of this point, the sediment moves southward for 16.5 km (10.25 miles) along the island complex (Stone and Ellis in press). The magnitude of predicted sediment transport is greater in the southern cell than the northern cell and is attributed to an increase in breaker angle and not breaker height. This finding underscores the control that wave refraction exerts on longshore transport trends along the island. A maximum rate of 88,000 m³/yr (~120,000 yd³/yr) is calculated in the southern cell and 66,000 m³/yr (~90,000 yd³/yr) in the northern cell.

2.9.5 Barataria Bay

For the Barataria Bay coastal segment estimates of longshore transport are based on shoreline erosional-depositional trends. Approximately 146,000 m³/yr (~200,000 yd³/yr) is calculated to move eastward along Grand Isle. Similarly, an estimated 85,000 yd³/yr of sediment moves westward from Sandy Point toward Barataria Bay. The Caminada Moreau headland contains a drift divide mid-way along the headland.

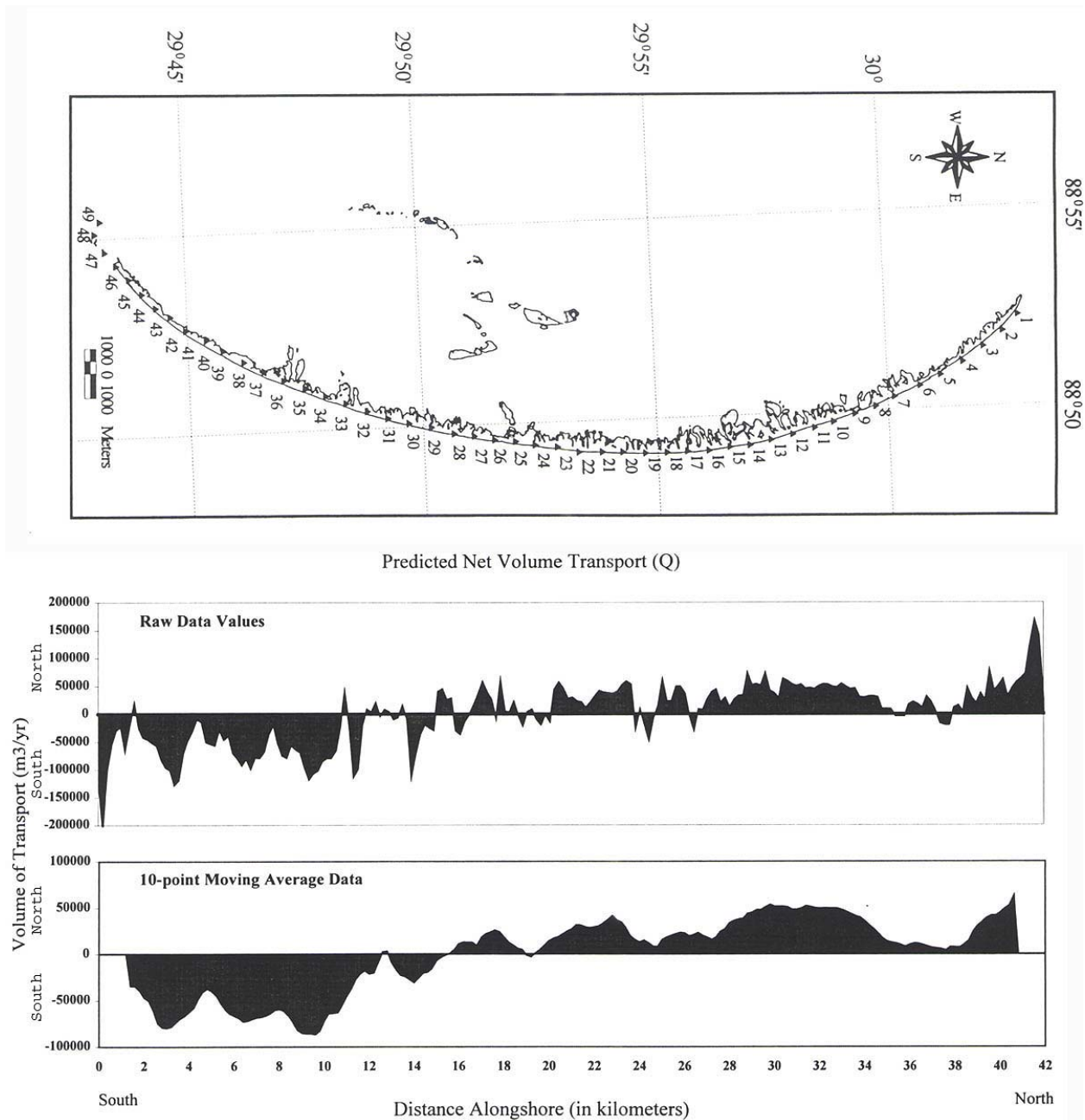


Figure D.2-25. Longshore sediment transport estimates for the Chandeleur Islands using numerical wave modeling (from Ellis 1998).

2.10 Cross-Shore Sediment Dispersal

Cross-shore sediment transport is the movement of sediment in a direction perpendicular to the shoreline. Rates of cross-shore sediment transport are difficult to quantify by direct measurements. As a result, little is known about the dynamics of this process. Cross-shore movement of sediment includes the sand that is eroded from the beach and transported offshore during storms, as well as the sand moved onshore by the process of overwash or during post-

storm recovery by fairweather waves. At the same time, storm waves breaking over low barriers wash sand into backbarrier marshes. This process provides a mechanism for the barrier islands to migrate landward and to reestablish sand platforms that are colonized by marsh vegetation. Another important process of cross-shore sediment transport occurs during the passage of cold-fronts in the vicinity of the Atchafalaya Delta. The strong offshore winds that accompany cold-fronts resuspend mud from the bottom that was recently deposited by the Atchafalaya River. Wind stresses and river discharge produce seaward directed currents that transport the suspended clays to the outer shelf (Roberts 2003).

The measurement and documentation of onshore sediment transport is a complex process that is not well understood. However, some examples do occur along the Louisiana coast indicating that this phenomenon is taking place. For example, the accumulation of sand in the lee of several of the breakwaters at the eastern end of Raccoon Island (Figure D.2-26). Bottom boundary layer and sediment transport measurements made on the shoal indicate a net onshore mean current and sediment flux during both fair-weather and storm conditions. Sand not only is deposited in the lee of the structures but has, as shown in the figure, accumulated between the breakwater gaps and seaward of them.



Figure D.2-26. Oblique aerial of Raccoon Island looking west (Photograph taken by G.W. Stone in 2000).

Another example of the transfer of sediment on the inner continental shelf has been documented in the region offshore of the eastern delta plain (List et al. 1989). Bathymetric change maps for the area seaward of the Bayou Lafourche shoreline eastward to the Plaquemines Delta from 1878-1989 reveal a pattern of large-scale sediment redistribution on the inner shelf

(Figure D.2-27). As seen in Figure D.2-27, since 1878 there has been widespread erosion in the vicinity of the Bayou Lafourche headland and the Plaquemines delta region seaward to the 15 m (50 ft) isobath. The greatest amount of erosion occurred closest to shore (> 2.5 m; 8.2 ft) and decreased seaward. In contrast, the embayments fronting Barataria and Terrebonne Bays have been depositional over the same time period, with more than 2.5 m (8.2 ft) of vertical accretion in certain areas (Figure D.2-27). The erosional and depositional trends suggest that there has been transport of sediment from headlands toward the intervening embayments. The pathway and mechanisms of sediment erosion, transport, and deposition are not well understood. Shoreline and nearshore promontories are sites of wave focusing and erosion, whereas embayments are commonly areas of wave dispersal and sedimentation.

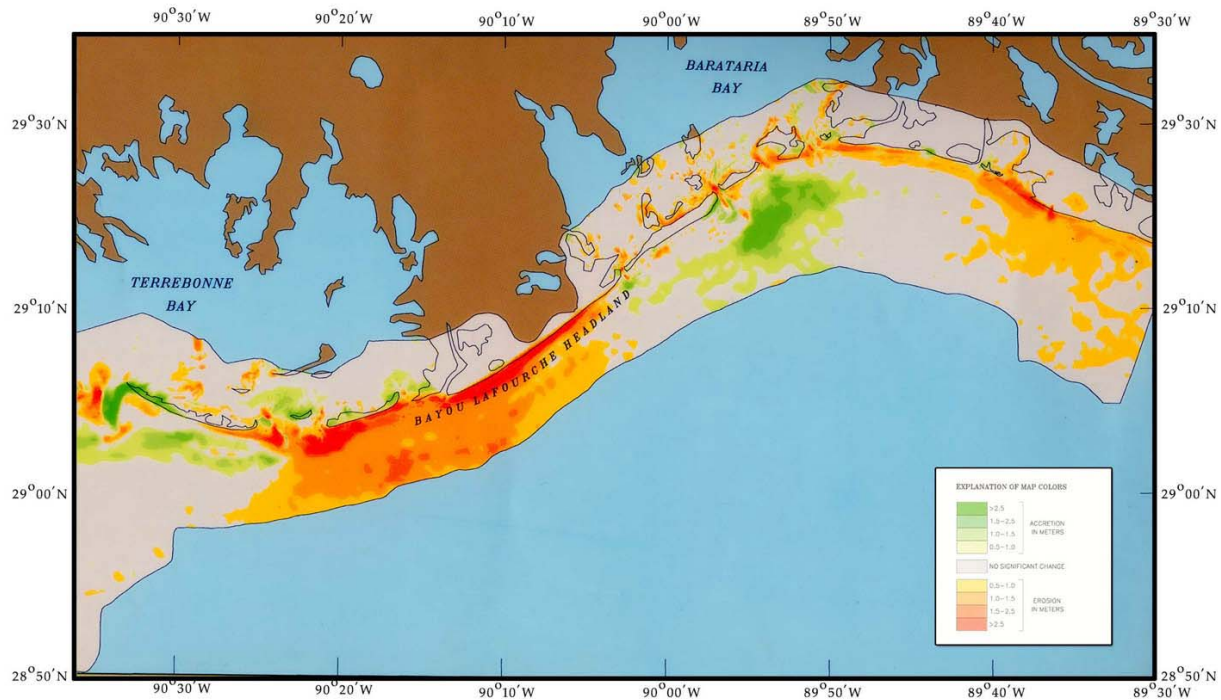


Figure D.2-27. Sea floor changes from 1878 to 1989 in coastal Louisiana (from List et al. 1989).